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Contourites of the Gulf of Cadiz: A high-resolution record of the paleocirculation of the Mediterranean outflow water during the last 50,000 years

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Abstract

The Mediterranean outflow water (MOW) paleocirculation during the last 50,000 years has been inferred from the grain-size distribution of contourite beds in core MD99-2341 from the Gulf of Cadiz (Southern Iberian Margin–Atlantic Ocean). Three main contourite facies are described. Their vertical succession defines two contourite sequences that reveal past variations of the MOW bottom-current velocity. A comparison of contourite sequences and the planktonic $\delta^{18}O$ record of core MD99-2341 with the $\delta^{18}O$ record from Greenland Ice Core GISP2 show a close correlation of sea-surface water conditions and deep-sea contouritic sedimentation in the Gulf of Cadiz with Northern Hemisphere climate variability on millennial timescales. A high MOW velocity prevailed during Dansgaard-Oeschger stadials, Heinrich events and the Younger Dryas cold climatic interval. The MOW velocity was comparatively low during the warm Dansgaard-Oeschger interstadials, Bølling-Allerød and the Early Holocene. Rapid sea-level fluctuations on the order of 35 m during Marine Oxygen Isotope Stage 3 are considered to have exerted limiting controls on the MOW volume transport and thus positively modulated the MOW behaviour during the last 50 kyr.

Keywords: Gulf of Cadiz; Contourites; Mediterranean outflow water; Paleoclimate changes; Sea-level fluctuations; Paleocirculation

1. Introduction

The last glacial period was marked by rapid climatic oscillations described in the records of Greenland ice cores (Dansgaard et al., 1993; Grootes et al., 1993). These Northern Hemisphere climatic oscillations, named Dansgaard-Oeschger (D-O) stadials and interstadials (cold and warm phases, respectively) are also recorded in ocean sediments (Bond et al., 1993). During some of the coldest stadials, the deposition of ice-rafted detritus indicates that massive iceberg discharges (Heinrich events – HE) occurred in the North Atlantic (Heinrich, 1988; Bond et al., 1992). These 'Heinrich layers' have been recognized as far south as the Portuguese margin (Schönfeld, 1993; Lebreiro et al., 1996; Baas et al., 1997; Bard et al., 2000; Schönfeld and Zahn, 2000; de Abreu et al., 2003) and in the Gulf of Cadiz (Reguera, 2001; Llave et al., 2006; Voelker et al., 2006). Meltwater

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discharge associated with these ice surges induced rapid changes in North Atlantic thermohaline circulation (Broecker et al., 1990; Rasmussen et al., 1997; Elliot et al., 2002). The atmospheric imprint of associated climatic oscillations modified the water-mass exchanges between Atlantic Ocean and Mediterranean Sea (Sierro et al., 2005).

The Gulf of Cadiz is located in the eastern Atlantic Ocean, west of Gibraltar Strait (Fig. 1). Present-day circulation is dominated there by antagonistic currents: at the sea surface, the North Atlantic Surface Water and the North Atlantic Central Water (NACW) flow into the Mediterranean Sea while a deep undercurrent, named the Mediterranean outflow water (MOW) flows out of the Mediterranean Sea (Ambar and Howe, 1979; Ambar, 1983). The MOW is a warm (13 °C) and saline (38‰) water mass and its dynamics are controlled by the integrated evaporative balance of the Mediterranean Sea (Bryden and Stommel, 1982) and more precisely by the interannual to decadal variability of its two sources (Astraldi et al., 2002): The Levantine Intermediate Water (LIW) is formed near Rhodes, and the Western Mediterranean Deep Water (WMDW) which is formed in the Gulf of Lions (South of France), especially during cold and windy winters (Lacombe et al., 1985; Rohling et al., 1998). The WMDW contributes only 10% to the MOW flow today (Kinder and Parilla, 1987). However, Voelker et al. (2006) consider the WMDW as a substantial part of the MOW during the last glacial period. They argued that buoyancy loss of glacial LIW could have reduced the density gradient between Mediterranean intermediate and deep waters and thus led to an increased contribution of WMDW to the MOW (Bryden and Stommel, 1984; Voelker et al., 2006). After exiting Gibraltar Strait, the MOW mixes with NACW in the Gulf of Cadiz (Baringer and Price, 1999) and moves northwest along the western Iberian Margin. The MOW divides into two major core layers: the Mediterranean Upper Water (MU, Fig.1) and the Mediterranean Lower Water (ML, Fig.1). The MU is centred between 400 and 600 m. The density of this core layer is in disequilibrium with the ambient NACW, and it is held at that shallow depth only by the Coriolis force induced by its high flow velocity. The slightly more saline and slower-moving

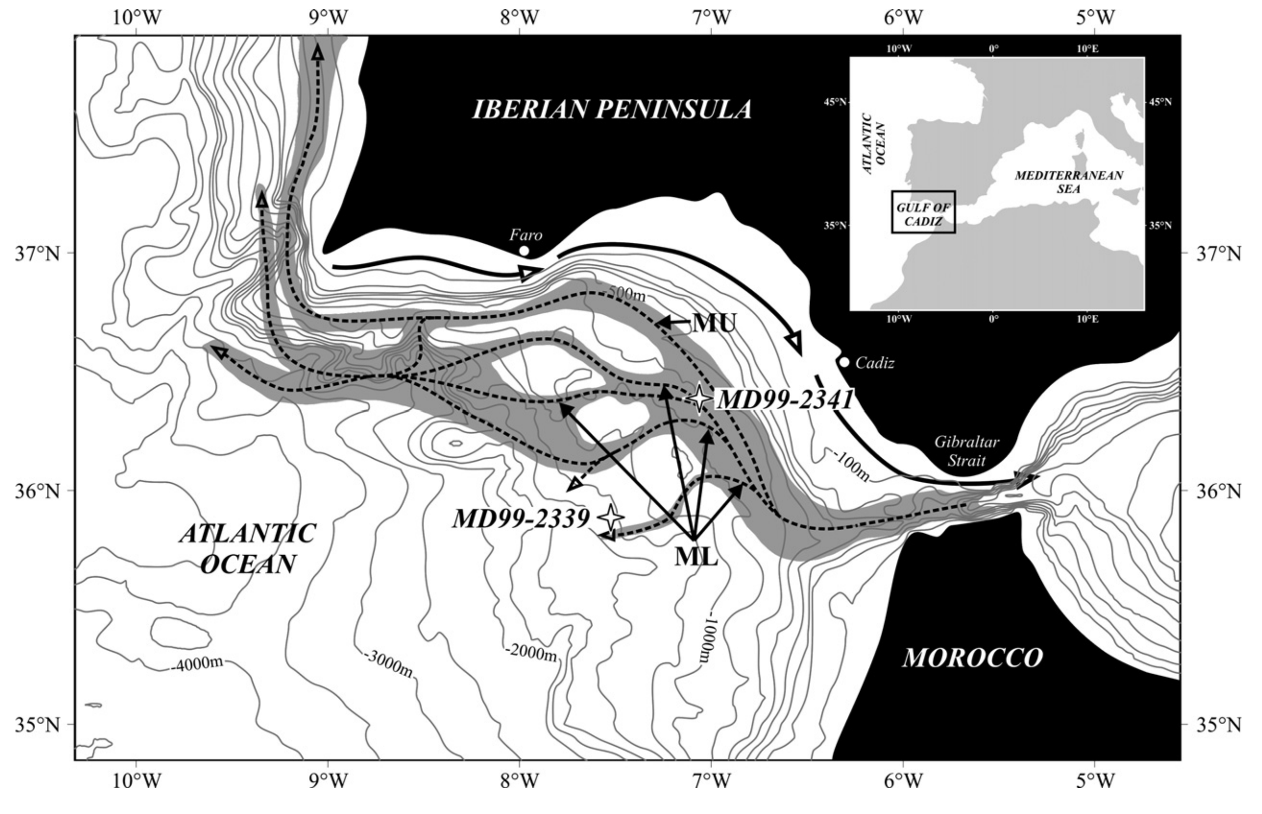


Fig. 1. General map of the Gulf of Cadiz showing the general circulation of the MOW and location of cores MD99-2341 (this study) and MD99-2339 (Voelker et al., 2006). Dotted arrows represent the Mediterranean outflow water (MOW) pathway, composed of the Mediterranean Upper Water (MU) and the Mediterranean Lower Water (ML). Continuous arrows in the northern part of the Gulf of Cadiz represent the Atlantic inflow pathway.

ML spreads between 600 and 1200 m (Zenk and Armi, 1990; Baringer, 1993; Bower et al., 1997). At this depth, the MOW loses contact with the seafloor in places and spreads over the North Atlantic Deep Water, before continuing westward and northward into the North Atlantic (Iorga and Lozier, 1999).

The impact of the MOW on sediment distribution and dynamics in the Gulf of Cadiz is depicted by contourite deposits (Gonthier et al., 1984). They build up hectometre thick and kilometre long sediment drifts (Kenyon and Belderson, 1973; Faugères et al., 1985). The drift sediment bodies and the average grain-size of surface sediments mirror the variability of MOW hydraulic energy (Kenyon and Belderson, 1973; Gardner and Kidd, 1987; Nelson et al., 1993; Nelson et al., 1999; Hernández-Molina et al., 2003; Mulder et al., 2003). Previous studies suggest that vertical variations of average grain-size in individual contourite beds are records of bottom-current variations (Gonthier et al., 1984; Faugères et al., 1986; McCave et al., 1995a; Llave et al., 2006; Voelker et al., 2006). Consequently, coarse-grained contourites are interpreted as deposits of increased MOW velocity in the Gulf of Cadiz (Gonthier et al., 1984; Faugères et al., 1985; Faugères et al., 1986; Nelson et al., 1993; Mulder et al., 2002; Llave et al., 2006; Voelker et al., 2006). Three periods of enhanced MOW flow have been inferred since the Last Glacial Maximum. They are named as Peak Contourites I, II and III, and have been dated to 14–15 ka Before Present (BP), 10–11 ka BP (Younger Dryas and Termination 1B) and 3 ka BP respectively on the radiocarbon scale (Faugères et al., 1986; Vergnaud-Grazzini et al., 1989). The coarse-grained Peak Contourites I to III are associated with high benthic $\delta^{13}C$ values (Vergnaud-Grazzini et al., 1989), high (smectite + kaolinite)/(illite + chlorite) ratios (Grousset et al., 1988), abundant benthic foraminifers with a high diversity (Caralp, 1988), and bottom-current sensitive, elevated epibenthic species (Schönfeld and Zahn, 2000). On the other hand, a lower MOW intensity is observed during intervals between 11–14 ka BP (Bølling-Allerød) and 5–9 ka BP (Early Holocene) on the radiocarbon scale. These periods are characterized by fine-grained contourite deposition, low benthic $\delta^{13}C$ values (Vergnaud-Grazzini et al., 1989), low (smectite + kaolinite)/(illite + chlorite) ratios (Grousset et al., 1988), and low abundances of elevated epibenthic foraminifers (Schönfeld, 2002). The difference in grain-size, which is even visually recognisable, therefore appears as a distinctive indicator of the MOW intensity. The relationship between grain-size variations with cold and warm climatic intervals suggests a strong correlation between climatic changes and MOW activity. This relationship during the last glacial period and in particular during D-O oscillations has recently been described for the central and western Gulf of Cadiz (Llave et al., 2006; Voelker et al., 2006).

In this paper, we present paleoclimatological and sedimentological records from a long piston core collected in the eastern part of the Gulf of Cadiz. Our aim is to relate high-frequency variations of the MOW to climate variability by defining sedimentary facies and small-scale sequences in the context of a high-resolution stratigraphic framework. We present a significant correlation between changes in the MOW intensity and D-O cycles and Heinrich events during the last 50 kyr. In a new integrated approach, rapid sea-level oscillations associated with millennial-scale climate fluctuations during the last glacial period are also considered.

2. Methods

This study is based on a 19.5 m long piston core collected northwest of the Strait of Gibraltar (MD99-2341; 36°23′25″N/07°3′94″W; 582 m water depth) by the *R/V Marion Dufresne* during the IMAGES V/GINNA cruise (1999). The core location is the south-eastern part of the Faro-Cadiz sheeted drift, north of the Huelva contourite channel, and at present directly under the influence of the MOW (Hernández-Molina et al., 2006; Llave et al., 2006) (Fig. 1). The core was opened, described, and sampled every 5 cm for oxygen isotopes and sedimentological studies at GEOMAR (Kiel, Germany). The samples were freeze-dried, weighed, and washed through a 63 μm sieve. The residues were dried, weighed, and the 150 μm fraction was separated for the planktonic foraminiferal census. The >150 μm fraction was also split to obtain an aliquot of approximately 500 specimens of planktonic foraminifera. We concentrated on the proportion of *Neogloboquadrina pachyderma* (sinistral) given as percentages. In the aim to display the presence of possible ice-rafted debris, lithic particles were counted in the fraction >250 μm.

Stable oxygen and carbon isotope measurements were made on 8 to 15 specimens of planktonic foraminifera *Globigerina bulloides* from the >250 μm size fraction. Measurements were made in the isotope laboratory at Bremen University with a CARBO KIEL automated carbonate preparation devices linked on-line to a FINNIGAN MAT 252 mass spectrometer. Long-term reproducibility was 0.08‰ for $\delta^{18}O$ as calculated from replicate analyses of the internal carbonate standard (Solnhofen Limestone).

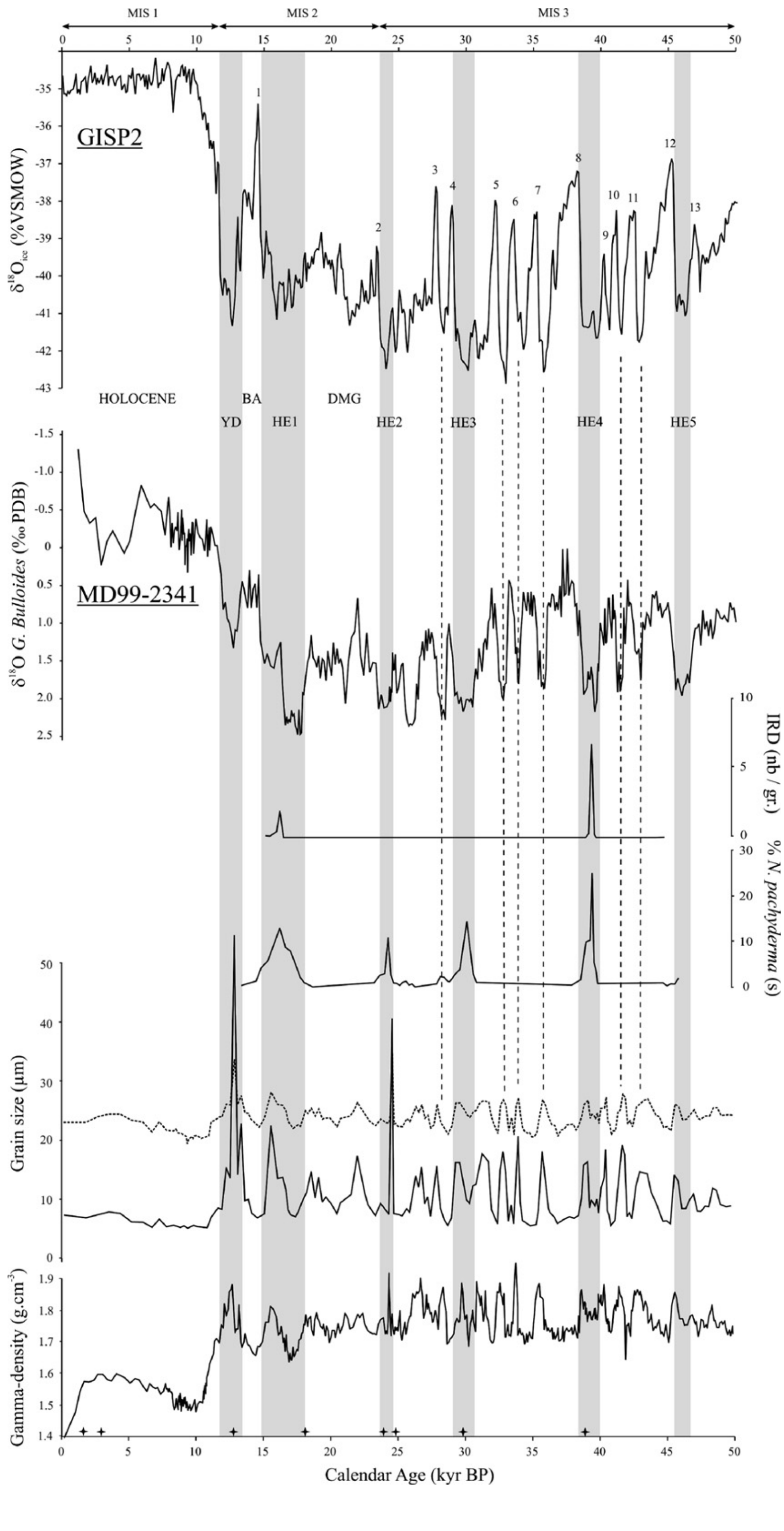
Table 1
Age-model for core MD99-2341 (Llave et al., 2006)

Depth (cm)	^{14}C Lab. number	^{14}C age (ka BP)	1 sigma error (year)	Correlation with core	Depth therein (cm)	^{14}C age [a] (ka BP)	Calibrated age (cal. ka)	Remarks
0							0.000	Core top
5	KIA14636	1.585	25				1.143	
25				M39008-3	8	2.660	2.863	
65	KIA14637	5.845	35				6.257	
95				M39008-3	68	6.954	7.874	
151				M39008-3	112	7.901	8.844	
205				M39008-3	172	8.320	9.283	
255	KIA14638	9.120	50				9.732	
285				M39008-3	332	9.450	10.717	
315				M39008-3	347	9.686	11.017	
370	KIA14639	11.130	50				12.689	
395				M39008-3	418	11.348	13.344	
445				SU81-18	250	12.460	14.705	
475				M39008-3	463	13.540	16.201	H1
485	KIA14640	14.210	80				16.465	
565	KIA14641	15.010	110					Discarded [b]
580	KIA14642	15.720	100				18.206	
675.9				GISP2	193,883		20.832	c/w transition
732.5				GISP2	198,598		23.405	Trans. to IS2
755	KIA14643	20.940	130				24.257	
805	KIA14644	21.530	190				24.846	
937.5				GISP2	205,577		27.832	Trans. to IS3
980.2				GISP2	207,552		29.011	Trans. to IS4
1005	KIA14645	26.290	240				29.776	
1111.9				GISP2	212,681		32.296	Trans. to IS5
1176.6				GISP2	214,647		33.618	Trans. to IS6
1275.8				GISP2	217,689		35.273	Trans. to IS7
1285	KIA14646	32.040	560					Discarded [c]
1412.8				GISP2	223,121		38.388	Trans. to IS8
1435	KIA14647	33.250	570				38.85	Diff. corr. [d]
1470				MD952039	1020	34.150	39.379	H4
1602.8				GISP2	227,549		41.151	Trans. to IS10
1678.8				GISP2	230,290		42.529	Trans. to IS11
1777.3				GISP2	235,822		45.371	Trans. to IS11

The chronostratigraphic framework was further improved by correlation of the oxygen isotope curve with the parallel-core M39008-3 (Cacho et al., 2001) in the upper part of the core and with the GISP2 ice core δ^{18}O record in the lower part of the core.
The data provided by Voelker et al. (2000) suggest a correction of 5600 years.
[a] Reservoir correction of 400 years subtracted.
[b] Radiocarbon age is too young possibly due to recrystallization of pteropod shell.
[c] Radiocarbon age is too young, possibly due to contamination with tap water precipitates.
[d] This dating is in the range of low geomagnetic intensities and high age offsets around IS8.

Radiocarbon datings were performed on 12 samples from core MD99-2341 (Table 1). Radiocarbon ages were determined via accelerator mass spectrometry (AMS) at the Leibniz-Labor, Kiel University. The precision ranges from ±25 to ±570 years (S.D.). ^{14}C ages younger than 20 kyr were calibrated to calendar years by using the web-based Calib 4.3 program (Stuiver and Reimer, 1993) and an ocean age reservoir correction of 400 years. For older datings, the correction provided by Laj et al. (1996) and Voelker et al. (2000) was used the correction provided by Laj et al. (1996) and Voelker et al. (2000).

The sedimentological study and the determination of sedimentary facies and sequences are based on visual description, gamma-density measurements (logged on board with a *Geotek Multisensor Core Logger*) and grain-size analyses, performed with a Malvern MASTERSIZER (University of Bordeaux 1, France), using median grain-size (D_{50}) and the 10–63 μm fraction (McCave et al., 1995b). Because D_{50} oscillations are more significant while showing congruent fluctuations with mean 10–63 μm in core MD99-2341, we use this grain-size parameter throughout this study.



MIS 1
MIS 2
MIS 3
GISP2
δ18O ice (%VSMOW)
HOLOCENE
BA
DMG
YD
HE1
HE2
HE3
HE4
HE5
MD99-2341
δ18O G. Bulloides (‰ PDB)
IRD (nb / gr.)
% N. pachyderma (s)
Grain size (μm)
Gamma-density (g.cm-3)
Calendar Age (kyr BP)

3. Results

3.1. Chronostratigraphic framework

The chronostratigraphy of core MD99-2341 is primarily based on 12 ^{14}C-AMS datings and the recognition of Heinrich events (Llave et al., 2006). Two datings were discarded: KIA14641 at 565 cm was performed on only one pteropod specimen that shows an age slightly younger than inferred by the datings above and below. This might be affected by tap water contamination or recrystallisation. Dating KIA14646 at 1285 cm is apparently also too young, only the lower 1-sigma error range fits the adjacent GISP2 core correlation points. As this sample had only a small amount of carbonate material, a substantial tap water contamination may be considered.

HE 1 through 5 were characterised in core MD99-2341 by the occurrence of *N. pachyderma* (s) (Fig. 2). Percentages reach 2% to 20% of total planktonic foraminifera with a maximum of 25% in the level of HE 4. *N. pachyderma* (s) is usually found in the North Atlantic Ocean during Heinrich events, reaching up to 100% of total foraminifera in high latitudes (Bond et al., 1993) and 25–50% off southern Portugal (Lebreiro et al., 1996; Cayre et al., 1999). HE 1 and HE 4 were also associated with coarse lithic particles of dolomite, feldspar, basalt, and hematite coated quartz grains (2 and 14 grains per gram of dry sediment in the fraction >250 μm) (Llave et al., 2006). Theses particles are consequently considered to be ice-rafted debris derived from the Laurentide Ice Sheet (Baas et al., 1997; Llave et al., 2006; Voelker et al., 2006). The concentrations are an order of magnitude lower than in sediment cores off northern and southern Portugal (Baas et al., 1997; Schönfeld and Zahn, 2000; Schönfeld et al., 2003). The ice-rafted debris maximum is coeval with the maximum abundance of *N. pachyderma* (s) during HE 1 and HE 4.

The chronostratigraphy provided by radiocarbon datings and Heinrich events facilitated a paleoenvironmental interpretation of the planktonic δ^{18}O record of *G. bulloides* in core MD99-2341 (Mulder et al., 2002; Llave et al., 2006). δ^{18}O values exhibits the typical Holocene to Glacial increase of about 1.5‰ on average and range from −1.10‰ to 2.68‰ (Fig. 2). These values are in agreement with the nearby core MD99-2339 (Voelker et al., 2006) which indicates that no persistent hydrographic boundary in surface waters prevailed between both core locations during the late Pleistocene and Holocene. The heaviest δ^{18}O values are recorded at a precursor event to HE 2 and at the beginning of HE 1. This level was previously denoted as the Last Isotopic Maximum (Schönfeld et al., 2003). The record shows superimposed millennial-scale variability during Termination I and in the earlier part of the record. The short-term variability mirrors D-O cycles as observed in Greenland ice cores (Dansgaard et al., 1993; Meese et al., 1997; Cacho et al., 2001) and exhibits depletions and enrichments of ~1.5‰ associated with the D-O Interstadials and Stadials, respectively. As a result, the chronostratigraphic framework of core MD99-2341 was improved by correlation of the planktonic oxygen isotope record with the GISP2 ice core record. Midpoints of cold–warm transitions at the beginning of D-O Interstadials in core MD99-2341 were correlated with equivalent structures of the GISP2 record (Shackleton et al., 2000). Finally, a close correlation of the oxygen isotope curve with the parallel gravity core M39008-3 (Cacho et al., 2001) provides additional age control points for the upper 395 cm. The comparison of the oxygen isotope curves and AMS datings from both cores, which are only 1.15 km apart, revealed lateral variations in sediment thickness of only a few tens of centimetres. As such, no stretching of the uppermost meters by the Calypso device is recognised for core MD99-2341 (Löwemark et al., 2006).

The age model reveals that core MD99-2341 extends back to 50.1 kyr, and thus comprises Marine Oxygen Isotope Stages (MIS) 1, 2 and most of MIS 3 (Fig. 2). Correlation of the GISP2 δ^{18}O record with the planktonic oxygen isotope curve of core MD99-2341 gives a correlation coefficient of $r^2 = 0.8$. This excellent correlation demonstrates a close and efficient teleconnection between the climate over Greenland and the oceanographic conditions in the Gulf of Cadiz over the last 50 kyr.

The sedimentation rates show an intriguing threefold division. The uppermost 95 cm of core MD99-2341, i.e.

Fig. 2. Correlations between paleoclimate and sedimentological records of core MD99-2341 from the Gulf of Cadiz and the Greenland Ice Core GISP2 $\delta^{18}O_{ice}$ record (Grootes et al., 1993; Meese et al., 1997) over the last 50 kyr. Numbers indicate Dansgaard-Oeschger interstadials 1–13 in the GISP2 record (Dansgaard et al., 1993). The planktonic δ^{18}O record of core MD99-2341 was measured in *G. bulloides* (Mulder et al., 2002; Llave et al., 2006). Associated to the δ^{18}O records, the abundance of ice-rafted debris in the fraction >250 μm and relative abundance (%) of planktonic foraminifera *N. pachyderma* (s.) displayed Heinrich event 1–5 (HE1–HE5, also marked by gray squares). Gamma-density and grain-size records of core MD99-2341 (continuous line corresponds to D50 (μm) and dotted line corresponds to mean 10–63 μm) show relevant fluctuations, interpreted as variations of strength of the MOW, which are connected with the GISP2 records (vertical dashed lines). Marine Isotopic Stages (MIS) 1 to 3 and Holocene, Younger Dryas (YD), Bølling-Allerød (BA) and Last Glacial Maximum (LGM) periods are noted on top to ease the reading. Black stars display the position of AMS ^{14}C ages realized on core MD99-2341.

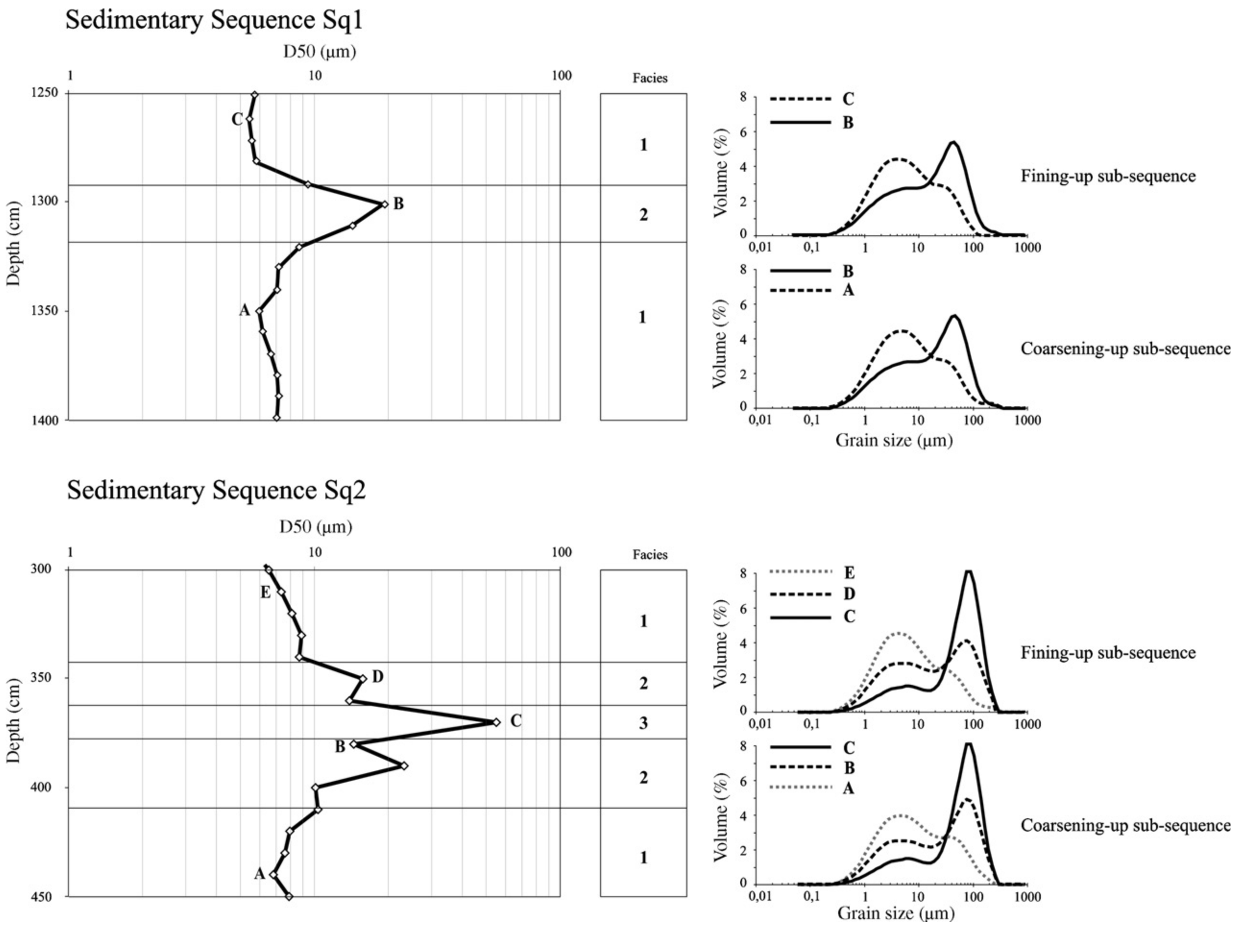


Fig. 3. Detailed sedimentological description of the sedimentary sequences Sq1 and Sq2.

the Late Holocene, show sedimentation rates of 12.1 cm kyr^{-1}. The Early Holocene from 95 to 315 cm shows very high sedimentation rates of 75.5 cm kyr^{-1} with peak values of up to 122.9 cm kyr^{-1}. The sedimentation rates, at 43.1 cm kyr^{-1}, are substantially lower during Termination I, MIS 2 and 3. There is a certain short-term variability of ±16.3 cm kyr^{-1} (1-sigma), which may due to enhanced winnowing during cold climatic intervals and stronger MOW current activity (Llave et al., 2006; Voelker et al., 2006). However, the resolution of the age model does not allow a precise discrimination of sedimentation rates during individual D-O interstadials and stadials in order to further constrain and quantify this relationship.

3.2. Sedimentary facies

A detailed analysis of sedimentological facies and sequences reveals that three main sedimentary facies were recognized in core MD99-2341 (Fig. 3).

Facies 1: silty-clay is the finest-grained and the most abundant facies. It is composed of structureless brownish grey sediment. D_{50} is generally less than 10 µm and the histogram shows an unimodal (mode = 6 µm) to bimodal (modes = 6 and ~30 µm) distribution. Visible bioturbation is low to moderate and often marked by thin filaments of iron sulphides interpreted such as *Trichichnus* burrows (Löwemark, personal communication). *Zoophycos* and *Chondrites* burrows are also observed. This facies prevails in core MD99-2341 for the Early Holocene and MIS 3, particularly during D-O interstadials.

Facies 2: clayey-silt forms decimetric to metric intervals of olive grey to brownish grey sediment. D_{50} is generally less than 15 µm and the histogram shows a bimodal distribution with modes ranging from 5 to 7 µm and from 30 to 80 µm. This facies can appear as ”mottled” i.e. containing pockets or lenses of either mud or silt in a finer matrix with gradual contacts (Gonthier et al., 1984). Bioturbation and burrowing are frequent in Facies 2. This facies occurs during HE 1, HE 3, HE 4, HE 5 and all D-O stadials.

Facies 3: clayey-sand represents the coarsest sediment that is recognised in core MD99-2341. It appears as irregular centimetre-thick beds. This facies is mostly surrounded by the mottled Facies 2. D_{50} ranges from 20 to 60 µm and the D90 reaches 130 µm. The histogram shows a bimodal frequency distribution with a principal mode between 60 and 100 µm. Facies 3 is recorded for the Younger Dryas and HE 2.

3.3. Sedimentary sequences

The gradual facies succession along the core MD99-2341 permits the definition of two sequences Sq1 and Sq2 (Fig. 3) both composed of a coarsening-up (silty-clay to clayey-silt or very fine clayey sand) and a fining-up (clayey-silt or very fine clayey sand to silty-clay) sub-sequence. The coarsening-up sub-sequence is usually thicker (30 to 80 cm) than the fining-up sub-sequence (10 to 30 cm). The two sequences are composed of the vertical succession of the facies 1-2-1 (Sq1) or 1-2-3-2-1 (Sq2). We interpret these vertical sequences as contourites formed by the MOW in this part of the Gulf of Cadiz (Faugères et al., 1984; Gonthier et al., 1984). Grain-size variations along the whole vertical sequences Sq1 and Sq2 suggest a progressive increase and then decrease of bottom-current competency and velocity. Progressive contacts between facies, i.e. 1–2 or 2–3, indicate quasi-steady currents. The reworking by bottom-currents is also displayed by the variation of the grain-size mode values along with contourite deposition as recorded in core MD99-2341 (Fig. 3). The sequence limits, interpreted as low bottom-current activity, show a main mode focussed around 5–7 µm. During speeding-up phases of bottom-currents and the settling of the coarsening-up sub-sequence, a second, coarser maximum around 55 µm becomes gradually the main mode in the grain-size distributions while the minimum at 5–7 µm diminished. The expression of the coarser mode displays significant winnowing of fine particles during periods of enhanced bottom-current activity. In agreement with the deposit environment (Hernández-Molina et al., 2003; Hernández-Molina et al., 2006; Llave et al., 2006) and despite the above-mentioned bioturbation and burrowing capable of removing erosive contacts and dynamic structures, the important bed thickness of the coarsening-up sub-sequences and their well-defined succession of grain-size distributions indeed confirm the contouritic origin of these deposits and discard the possibility of eventual bioturbated turbidites.

4. Discussion

4.1. Significant imprint of millennial-scale variability in contourite deposits in the Gulf of Cadiz

Correlation of GISP2 $\delta^{18}O$ with the planktonic oxygen isotope record (*G. bulloides*) of core MD99-2341 demonstrates a close connection between climate over Greenland and environmental conditions in Gulf of Cadiz surface waters over the last 50 kyr. The D-O variability is clearly recorded during the last glacial period and was first identified in the Gulf of Cadiz by Mulder et al. (2002). Isotopic data for core MD99-2341 are accompanied by the record of the polar foraminifera *N. pachyderma* (s). Incursions of this species confirmed that severe cold conditions prevailed during those D-O stadials associated with Heinrich events. With an abundance of *N. pachyderma* (s) reaching 25% of total planktonic foraminifera, HE 4 reveals the strongest influence of subpolar surface waters during the last 50 kyr (Colmenero-Hidalgo et al., 2004). However, summer sea surface temperatures were, at 10 °C, not lower than during Heinrich events H1 and H5 (Voelker et al., 2006). That indeed icebergs drifted over the site of MD99-2341 is only documented for HE 1 and HE 4 by the presence of ice-rafted debris.

Chronostratigraphic framework and sedimentological data of core MD99-2341 show a clear relationship between millennial-scale climatic variations and contourite grain-size during the last 50 kyr. With a mean grain-size ranging from 5 to 55 µm, core MD99-2341 shows a repetitive succession of Sq1 and Sq2 sequences, also displayed in the density record. These sequences are strictly correlated with the D-O cycles during MIS 3. D-O interstadials correlate with fine-grained deposits (Facies 1), in particular interstadials 3 to 8 (Fig. 2). Conversely, D-O stadials and Heinrich events correlate with high grain-size mode values (Facies 2). D-O cycles are thus characterised by the succession of Sq1 sequences, except for HE 2 which is characterised by a Sq2 sequence. Alternations of successive fine- and coarse-grained contourites are also observed during the last 14 kyr. The Bølling-Allerød and the Early Holocene are characterized by fine-grained deposits (Facies 1), whereas the Younger Dryas shows coarse-grained deposits (Facies 3). Due to the presence of Facies 2 in the transition Bølling-Allerød/Younger Dryas and Younger Dryas/Early Holocene, the 14–5 kyr period displays a Sq2 sequence. Because contourite grain-size and density mirror bottom-current velocities, we suppose an enhanced MOW velocity at site MD99-2341 during the Younger Dryas, Heinrich events and D-O stadials. A sluggish MOW flow is inferred during Early Holocene, Bølling-Allerød and D-O interstadials.

The lowest grain-size values are recorded in core MD99-2341 during the Early Holocene and thus reveal the weakest activity of the MOW on the Faro-Cadiz sheeted drift during the last 50 kyr. An exceptionally fine-grained interval is also reported from other cores in the eastern and central Gulf of Cadiz at depths of both, Mediterranean Upper and Lower waters (Mulder et al., 2002; Voelker et al., 2006). These observations are consistent with an intermittent reduction in MOW flow between 9 and 6 ka at 600 to 900 m depths further downstream off southern Portugal (Schönfeld and Zahn, 2000). This flow reduction was synchronous with sapropel S1, when a stagnation of Mediterranean deep waters due to a stable surface water stratification led to a decrease of the Atlantic/Mediterranean water-mass exchange (Béthoux and Pierre, 1999). Huang et al. (1972) and Nelson et al. (1993) even invoke a current reversal in the Gibraltar Strait during this period. Numerous studies demonstrated, however, that the sense of the circulation does not change from its present antiestuarine mode since the Last Glacial Maximum (Grousset et al., 1988; Rohling and Bryden, 1994; Myers et al., 1998; Rohling, 1999; Rohling and De Rijk, 1999) and rather point to supporting evidence for a weakened MOW flow (Zahn et al., 1987; Vergnaud-Grazzini et al., 1989; Schönfeld and Zahn, 2000; Rogerson et al., 2005; Voelker et al., 2006). The Early Holocene very fine-grained deposits are therefore interpreted as hemipelagic sediments that were deposited under a sluggish MOW flow. Sedimentation rates in core MD99-2341 increased significantly during this period and reach 75.5 cm kyr^{-1} with peak values of up to 122.9 cm kyr^{-1}. This could be interpreted as a decrease of winnowing of fine particles. Contrary to the phases of stronger MOW flow, the fine suspension could settle and forms thick contourite deposits (Faugères et al., 1986). However, a substantial decrease in sedimentation rates of up to 8.5 cm kyr^{-1} is described for core MD99-2339 and interpreted as an intermittent reduction of both, flow strength and sediment delivery (Voelker et al., 2006). As core MD99-2341 is closer to the coast than core MD99-2339, the Early Holocene sedimentation rates could rather be influenced by sediment supply from the adjacent Iberian continent than solely by bottom-current activity. This interpretation is corroborated by a contemporary enhanced fluvial activity in Spain and hence riverine sediment supply at times of a warm and wet climate there (Thorndycraft and Benito, 2006). The coarsening-upwards trend in the uppermost interval of the core reveals a reactivation of MOW activity and Atlantic–Mediterranean water exchange during the Late Holocene.

4.2. Influence of climatic and sea-level changes on the MOW strength over the last 50 kyr

The spatial and temporal dynamics of the MOW are controlled by the integrated evaporative balance of the Mediterranean Sea and by variations of the intensity in Mediterranean deep-water formation (Bryden and Stommel, 1982). Climatic changes, sea-level and the changing hydraulic control conditions in the Strait of Gibraltar are the main factors which induced variations of the Mediterranean–Atlantic water-mass exchanges (Béthoux, 1984; Bryden and Stommel, 1984; Rohling and Gieskes, 1989; Rohling and Bryden, 1994; Matthiesen and Haines, 1998; Myers et al., 1998; Matthiesen and Haines, 2003). Because climate affects the deep-water formation in Mediterranean Sea and WMDW could well have been a major source of the MOW during the last glacial (Myers et al., 1998; Voelker et al., 2006), climatic changes over the western Mediterranean Sea have to be considered. The last glacial was characterized by millennial-scale climate fluctuations (Dansgaard et al., 1993; Grootes et al., 1993; Shackleton et al., 2000) and the western Mediterranean Sea was very sensitive to the rapid climatic variability in the glacial North Atlantic (Rohling et al., 1998; Cacho et al., 1999; Sierro et al., 2005). D-O cycles and Heinrich events in marine records of Alboran Sea and Menorca Margin have been associated with synchronous oscillations of sea surface temperatures and changes in Mediterranean deep water convection (Cacho et al., 2000; Sierro et al., 2005). Low sea-surface temperatures (Cacho et al., 1999), intense north-westerly winds, and dry, cool conditions on land (Combourieu-Nebout et al., 2002; Sanchez-Goni et al., 2002; Moreno et al., 2005) facilitated an increased formation of WMDW in the Gulf of Lions during D-O stadials and Heinrich events (Rohling et al., 1998; Cacho et al., 2000; Sierro et al., 2005; Voelker et al., 2006). The inflow of salt-enriched, subtropical Atlantic waters into the western Mediterranean Sea also increased surface water density and facilitated deep convection in the Gulf of Lions (Voelker et al., 2006). However, temporary incursions of subpolar surface waters during Heinrich events resulted in a brief slowdown of western Mediterranean deepwater overturn (Sierro et al., 2005). These injections of subpolar waters are depicted by lowerings of planktonic $\delta^{18}O$ (Combourieu-Nebout et al., 2002; Colmenero-Hidalgo et al., 2004; Sierro et al., 2005; Voelker et al., 2006). Core MD99-2341 shows temporary $\delta^{18}O$ depletions of 0.5‰, associated with ice-rafted debris, only during HE 1 and HE 4 however. This indicates that the Atlantic current flowed close to the coast and only occasionally affected the coring sites in the lower southern region of the Gulf of Cadiz.

The grain-size data for the last glacial period from core MD99-2341 show: (i) a stronger outflow during Northern Hemisphere coolings (Fig. 2); (ii) an acceleration of the MOW during HE 1 and HE 4 once the meltwater disappeared; (iii) a weaker outflow during D-O interstadials. This last result is in agreement with the reduction of Mediterranean deep convection (Cacho et al., 2000; Sierro et al., 2005) induced by increased precipitation associated with warmer temperatures over Southern Europe during D-O interstadials (Sanchez-Goni et al., 2002; Moreno et al., 2005).

If climatic conditions appear to be a prime component in the behaviour of the MOW, sea-level and ensuing hydraulic conditions in the Strait of Gibraltar have also to be addressed (Béthoux, 1984; Rohling and Bryden, 1994; Myers et al., 1998; Matthiesen and Haines, 2003; Rogerson et al., 2005). Low sea level during the last glacial caused a narrowed geometry of the Strait of Gibraltar and hence a strong decrease in the outflow volume (Béthoux, 1984; Bryden and Stommel, 1984; Rohling and Bryden, 1994; Matthiesen and Haines, 1998). The glacial outflow is estimated at 0.39 Sv compared with 0.86 Sv today. The restricted water exchange, lowered temperatures and a drier climate over the Mediterranean increased the salinity of the Mediterranean Sea (Bryden and Stommel, 1984; Rohling and Bryden, 1994; Rohling and De Rijk, 1999). Small-scale sea-level fluctuations of up to 35 m (between −100 and −63 m according to Siddall et al. (2003)) were associated with millennial-scale climate changes during MIS 3 (Yokoyama et al., 2001a,b; Cacho et al., 2002; Chappell, 2002; Siddall et al., 2003). These sea-level oscillations certainly also influenced the behaviour of the MOW although a precise stratigraphic relationship between D-O cycles and sea-level oscillations is not yet defined. Therefore it appears difficult to link sea-level oscillations with grain-size data of core MD99-2341 and MOW strength. Nevertheless, Chappell (2002) and more recently Jouet et al. (2006) in the western Mediterranean Sea demonstrated that sea-level highstands prevailed during D-O interstadials 8 and 12. More precisely, periods of sea-level rise occurred during HE 4 and HE 5. This pattern is consistent with the grain-size trend in core MD99-2341. HE 4 and HE 5 are characterized by high grain-size mode values while lower values characterized both following D-O interstadials. Both, cold climate conditions and relatively low sea level acted in the same direction and both effected an enhanced velocity of a more saline and thus denser MOW in the Gulf of Cadiz. As such the last glacial conditions promoted a more intensive ML (Rogerson et al., 2005; Llave et al., 2006).

Despite its shallower depth, the results of grain-size analysis from core MD99-2341 are, particularly for the MIS 3, in agreement with those of the deeper core MD99-2339 from the ML (Voelker et al., 2006). Numerous physical studies about the MOW demonstrated that the boundary between the Mediterranean Upper and Lower waters remained elusive. Since the associated core MD99-2341 and our results for the MU activity during the last glacial period are in disagreement both with model predictions about the Gibraltar exchange (Rohling and Bryden, 1994) and with numerous observations on the glacial MOW dynamics (Nelson et al., 1993; Schönfeld and Zahn, 2000; Rogerson et al., 2005; Llave et al., 2006), the observed velocity variability must be directly influenced by the ML (Llave et al., 2006). The assumption of a subordinate MOW branch, probably confined to the Huelva Channel, and passing through a region generally experiencing reduced flow during glacial times, is also not ruled out.

5. Conclusions

MOW paleocirculation in the Gulf of Cadiz has been inferred from the grain-size distribution of core MD99-2341. As displayed by the planktonic $\delta^{18}O$ record (*G. bulloides*), a high sedimentation rate allows climatic variability to be recorded in contourite deposits. A significant correlation of grain-size data and the planktonic $\delta^{18}O$ record with the $\delta^{18}O$ record from Greenland Ice Core GISP2 display a strong teleconnection between deep-sea contouritic sedimentation in the Gulf of Cadiz and Northern Hemisphere climate variability. Grain-size fluctuations are interpreted as changes in MOW intensity paced by climatic changes. A high MOW velocity prevailed during D-O stadial periods, Heinrich events and Younger Dryas while a low MOW velocity is inferred for D-O interstadial periods, Bølling-Allerød and Early Holocene. Cold climatic intervals, favourable to deep-sea water formation in Mediterranean Sea, thus intermittently increased the MOW intensity during the last 50 kyr. Associated with climate changes, the effects of rapid sea-level fluctuations during MIS 3 are also considered. While sea-level lowstands increased the salinity of the Mediterranean Sea and thus the density and the downslope velocity of the MOW, relative sea-level highstands occurred during D-O interstadials 8 and 12 seem to have reduced the bottom-current velocity. The progressive decrease of the MOW intensity was then associated with the preceding sea level rises, i.e. roughly coinciding with HE 4 and HE 5—interstadials transitions. The contemporary climatic transition, less favourable to deep-sea water formation in Mediterranean Sea, has been associated with effects of sea-level changes during these periods to reduce the MOW intensity at site MD99-2341. The integrated approach of climate and sea-level fluctuations appears consequently to be of primary importance to understanding MOW behaviour during the Holocene and Late Pleistocene.

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